Simple Snowdrift Model for Distributed Hydrological Modeling

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Abstract: A simple snowdrift model was developed, incorporated into a distributed winter-time hydrological model, and tested against snow measurements from a hillside in eastern Washington State. Snow movement can be an important factor in the distribution of spring soil moisture and runoff. Although current hydrological models often attempt to account for heterogeneities in precipitation distribution, they do not account for snowdrift. Snow melts and accumulates during the same times that it is redistributed. Therefore, evaluation required a snowmelt/accumulation model to be coupled with the snowdrift model. The snowmelt/accumulation model used the standard energy balance approach and performed well, i.e., standard errors of snow water equivalent $\approx$1 cm. The snowdrift model's simulated snow distribution generally agreed with observed snow distribution across a hill. Most notable were the model's ability to correctly place a snowdrift on the lee side of the hill and its ability to predict snow removal from nondrift areas. The effects of snow redistribution and the model's ability to reproduce these were obvious when overlaid on model results that ignored snowdrift.

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Introduction

At mid- to high latitudes and at high elevations, snowfall is a substantial precipitation form. In many systems, snowmelt is an important contributor to streamflow and groundwater recharge. There have been many studies and subsequent models developed that accurately simulate various snow processes: melt, snow densification, albedo dynamics, etc. (U.S. Army Corps of Engineers 1960; Schmidt 1972; Anderson 1976; Bengtsson 1980, Flerchinger and Saxton 1989; Grant 1992; Walter 1995; Tuteja and Cunnane 1997; Marks et al. 1999; Marks et al. 2001), but drift has received little attention. It is notable that although the impacts of heterogeneous rainfall distribution and snowmelt on soil moisture and hydrological responses have received much consideration (Ogden and Juliean 1993; Dunne 1998; Wood 1998), the impacts of snow drifting have received very little attention in hydrological modeling, despite the fact that there has been extensive research on snow drifting both with respect to its importance in water resources and drift process mechanics (Kind 1981; Steppuhn 1981; Steppuhn and McConkey 1988; Pomeroy and Gray 1990).

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Watershed hydrological models commonly represent wintertime-specific hydrological processes, snowdrift in particular, less mechanistically than other processes such as interflow, overland flow, and evapotranspiration (Young et al. 1989; Zollweg et al. 1996). Usually, distributed models simply accumulate snow when precipitation falls at air temperatures below freezing and model snowmelt using "degree day" (Hottelet et al. 1993; Rango and Martinec 1995) or bulk basin snowmelt (USACE 1960) relationships (Kustas et al. 1994; Ambroise et al. 1996; Cazorzi and Fontana 1996; Holko and Lepisto 1997; Lindstrom et al. 1997; Schreider et al. 1997; Frankenberger et al. 1999). The few distributed watershed models that consider mechanistic wintertime hydrological modeling limit the scope to snowmelt and snow accumulation (Abbott et al. 1986; Wigmosta et al. 1994; Flerchinger et al. 1996). This investigation develops a simple snowdrift model and demonstrates its application to a distributed hydrological model. Furthermore, this model was developed with "simplicity" as an objective to avoid overly complex parameterization that often plagues mechanistic, distributed hydrological models (Grayson et al. 1992), particularly those that simulate snow processes (Rango and Martinec 1995). It should be noted that there have been several recent snowdrift models developed independently of the hydrological modeling context that appear to successfully capture the fundamental processes but, again, are too complicated for easy assimilation with distributed hydrological models, especially models intended for water planning and management rather than research (Lykossov 2001; Lehning et al. 2000; Liston and Sturm 1998; Pomeroy et al. 1997; Sivardiere et al. 1995; Pomeroy et al. 1993; Gauer 2001).

Model Development

Heterogeneous snow distribution by wind may occur as snow falls or through saltation, or "ground drift," once the snow is already on the ground. This study's scope is limited to the latter. Snow will be entrained into the windstream if the windspeed

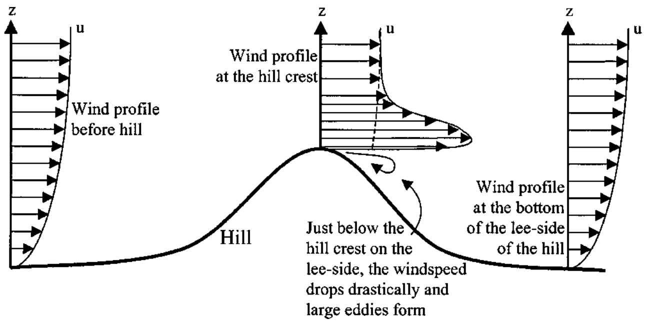

Fig. 1. Topographically influenced wind profiles

exceeds a threshold. The snow will leave the wind stream when the windspeed is substantially dissipated as occurs in the lee side of a hill or at the edge of heavy vegetation. Because snow's propensity to drift is largely density dependent, this model considers a two-layer snowpack with new snow overlaying old snow. For this model, only new snow is available to drifting; this is snow that has most recently fallen and is still light enough to be relatively easily moved by the wind.

Snow is entrained into the wind stream when the friction velocity U^* ($\mathrm{m \cdot s^{-1}}$) is greater than the snow's threshold shear velocity U^*_{th} ($\mathrm{m \cdot s^{-1}}$). The friction velocity can be expressed as (Campbell 1977)

$$U^* = \frac{ku}{\ln\left(\frac{z_u + z_0 - d}{z_0}\right)} \tag{1}$$

where k=von Karman's constant (0.41); u=windspeed ($\mathrm{m \cdot s^{-1}}$); z_u=the wind measurement height (meters); z_0=the surface roughness height (0.01 m for snow; McConkey 1992); and d=the zero plane displacement height [0 m for snow (Barry et al. 1990)]. The following expression for the snow's threshold shear velocity is based on limited data from Kind (1981):

$$U^*_{th} = 0.0195 + 0.021\sqrt{\rho_{ns}} \tag{2}$$

where ρ_{ns}=the density of new snow ($\mathrm{kg \cdot m^{-3}}$), i.e., the most recently accumulated snow. Once snow entrainment occurs, the snow flux J_s ($\mathrm{m^3 \cdot s^{-1}}$/m perpendicular to the wind direction), can be calculated by applying a simple force balance to the particles and utilizing an empirical proportionality constant (Kind 1981)

$$J_s = \frac{\rho_a U^*}{g\rho_{ns}}\left(0.25 + \frac{U_s}{3U^*_{th}}\right)\left[1 - \left(\frac{U^*_{th}}{U^*}\right)^2\right] \tag{3}$$

where ρ_a=the air density ($\sim$1.29 $\mathrm{kg \cdot m^{-3}}$); g=the acceleration of gravity (9.81 $\mathrm{m \cdot s^{-2}}$); and U_s=the terminal fall velocity of snow particles [0.75 $\mathrm{m \cdot s^{-1}}$; Kind (1981)]. Eq. (3) represents the carrying capacity of the wind; drift may be limited by the availability of new snow. Also, because blowing snow is highly susceptible to sublimation (Pomerory and Li 2000), when drifting occurs, this model assumes that the snow surface loses up to twice the amount calculated with Eq. (3) to account for direct losses to the vapor phase (Tabler 1975). Specifically, the amount of sublimated water mass removed from the new snow layer is equal to J_s or the balance of the new snow mass if J_s is greater than the remaining new snow.

Wind speed is influenced by topography, as shown in Fig. 1, and at any location, i, it can be expressed as

$$u_i = (C_\beta \Delta u_i + 1)u_0 \tag{4}$$

where u_0=the ambient windspeed below the ridge ($\mathrm{m \cdot s^{-1}}$); Δu_i=the change in windspeed at location i; and C_β=a unitless correction factor that accounts for wind striking a hill slope nonperpendicularly. Assuming an idealized, isolated ridge with triangular cross section, the change in windspeed at the crest or peak Δu_p can be approximated by $4\tan(\phi_g)$ (Stull 1988); ϕ_g is the ground slope.

On a regional or watershed scale ($<$50 km^2), locations on the highest ridges or hilltops typically have higher windspeed than lower elevations. One simple way to capture regional differences in windspeed is to scale by elevation. For a landscape or watershed characterized by a maximum elevation difference $Z_{\max}$ and an average ground slope ϕ_{avg}, the change in windpeed at relative elevation Z_i due to regional topography is approximated by

$$\Delta u_{\text{regional}} = 4\tan(\phi_{\text{avg}})\frac{Z_i}{Z_{\max}} \tag{5}$$

where Z_i is the local relative elevation. The writers currently have no good guidelines for how to define a "region" with respect to Eq. (5) and acknowledge that in some landscapes or over very large areas this could be nontrivial.

In addition to regional windspeed differences, local topography will also impact the windspeed at specific locations. The local change in windspeed at any windward location with an elemental length ΔL_i and local ground slope ϕ_i in a landscape can be approximated by

$$\Delta u_{\text{local}} = 4\tan(\phi_i)\left[\frac{\sin(\phi_{\text{avg}})}{Z_{\max}}\right]\Delta L_i \tag{6}$$

The total change in windspeed at any windward location in a landscape can be estimated as the sum of the regional, Eq. (5), and local, Eq. (6), windspeed changes.

Flow is turbulent and complicated on the lee side of a hill. Ideally, the windspeed is near zero just below the crest on the lee side and changes exponentially to its prehill condition at the bottom of the hill (Fig. 1). The following equation is used to describe the change in windspeed on the lee side of a hill:

$$\Delta u_{\text{lee}} = -\exp\left(\frac{4(Z_{\max} - Z_i)}{\sin(\phi_i)}\right) \tag{7}$$

To adjust the windspeed changes for hill slopes that are not perpendicular to the wind direction, the following correction factor can be used:

$$C_\beta = \cos(\beta_u - \beta_g) \tag{8}$$

where β_u is the wind direction from north, and β_g is the slope aspect from north.

This model considers $u=0$ immediately on the lee side of a hill, hedgerow, snow fence, etc., and on the windward side of a vegetated area. At the scale of distributed hydrological modeling and for our research sites, it was found that correctly predicting snow accumulation in conjunction with densely vegetated areas was relatively simple compared to topographical influences and, therefore, have not added any associated analysis in this paper.

In addition to windspeed, snow density is critical to snow drifting. This model uses a snow layer approach with new snow overlaying old snow. It is assumed that only new snow has a potential to move. Once the new snow layer's density ρ_{ns} exceeds 150 $\mathrm{kg \cdot m^{-3}}$, it is incorporated into the old snow layer. This density partition value is approximately the point at which metamorphic changes within the snowpack become density dependent (Anderson 1976). At $\rho_{ns} < 150\ \mathrm{kg \cdot m^{-3}}$, densification is attributed largely to the rounding of newly fallen snow crystals. At $\rho_{ns} > 150\ \mathrm{kg \cdot m^{-3}}$, densification is governed by overburden compaction. Snow densification and settling depends on the overload and snow temperature (Schroeter and Whitely 1987; Flerchinger and Saxton 1989; Barry et al. 1990; McConkey 1992). The present model updates daily snow density ρ_s using a method based on relationships used by Flerchinger and Saxton (1989) and McConkey (1992)

$$\rho_s = \rho_s^{t-1} + (\rho_{\max} - \rho_s^{t-1}) e^{-\mu} \tag{9}$$

$$\mu = 16[1 - \exp(T_{ns} - 1.5)]\exp\left(-\frac{2 d_{ns} \rho_{ns}^{t-1}}{3\rho_{\max}}\right), \quad \text{for new snow} \tag{10}$$

$$\mu = 16[1 - \exp(T_{os} - 1.5)]\exp\left[-\frac{d_{ns}\rho_{ns}^{t-1} + (2/3) d_{os}\rho_{os}^{t-1}}{\rho_{\max}}\right], \quad \text{for old snow} \tag{11}$$

where ρ_s^{t-1}=the snow density from the previous day; $\rho_{\max}$=a snow density maximum; and μ=the snow settling coefficient, which is a function of snow weight and temperature; d_{ns} and d_{os} are the depth of the new and old snow layers (in meters), respectively, and T_{ns} and T_{os}=the new and old snow temperatures (°C), respectively. The snow density maximum for new and old snow is 150 $\mathrm{kg \cdot m^{-3}}$ and 350 $\mathrm{kg \cdot m^{-3}}$, respectively (Schroeter and Whitely 1987). Each snow layer is considered homogenous, and the model calculates each layer's densification at its center of pressure.

Settling is enhanced when the snow is wet or when it is exposed to high winds. Densification and settling due to these processes can be complicated. We have simplified by assuming that when the snow's liquid moisture content is greater than zero or when more than half the new snow is removed by wind, the snow settling coefficient is doubled. Similarly, in places where drifting snow is accumulated, the snow tends to densify; therefore, accumulated drift snow is received at twice the new snow's density to account for this.

Methods

The snow drift model was incorporated into a more complete distributed Wintertime Hydrological Model (WHYM) (Walter 1995), and simulated results were compared with field measurements. Only the snow routines were enabled in WHYM in order to isolate the snow-related processes. Although many modelers continue to support watershed model performance based solely on how well the model reproduces stream hydrographs, hydrologists increasingly encourage corroboration of observed and predicted intracatchment hydrological processes for assessing the performance of distributed watershed models (Grayson et al. 1992; Wigmosta and Burges 1997; Wood 1998). It is difficult to quantitatively test a distributed model's ability to reproduce spatially varying processes over an entire landscape (Wigmosta et al. 1994), so the approach adopted in this study was to test the modeled snowdrift predictions against measured data from a hillside near Pullman in southeastern Washington State. Because it was not practical to measure snowdrift in the field without inadvertently including the effects of snow accumulation and melt, the routines for modeling these processes were retained in our hydrological model, and we subsequently evaluated our confidence in the model to simulate snowmelt and accumulation without drift.

The snow drift model was tested against data from a relatively isolated hill 10 km northeast of Pullman, Wash. (46.8° N, 117.2° W) collected over the winter of 1990–1991 by researchers at Washington State University. Snow first accumulated in December 1990 and lasted until January 1991. The elevation of the site is approximately 750 m above mean sea level (msl). The climate is subhumid with a cold winter season, and about 60% of the annual average precipitation (520 mm) falls between November and April. The mean annual temperature is 8.3°C.

Table 1. Pullman, Wash. Hill Slope Characteristics and Slope Positions for Modeled Cells

Slope aspect	Cells	Slope (%)	Hillside position[a]
South 160° from north	1–5	3	Foot
	6–8	7	Foot
	9–10	10	Foot
	11–14	14	Back
	16–19	18	Back
	20–21	19	Top
	22–24	9	Summit
North 30° from north	25–26	18	Top
	27–28	31	Back
	29–30	24	Back
	31–32	13	Foot
	33–35	6	Foot

[a]Position names are included for reference purposes in subsequent figures and tables.

The model was applied to the surrounding watershed using 10 m×10 m cells. The hill was represented by a straight transect of 35 cells as characterized in Table 1. The hill was part of a ridge with two distinct faces characterized by aspects of 160° from north (south side of hill) and 30° from north (north side of hill). The hill was located in a field that was partially fallow with conventionally tilled wheat stubble and partially planted in conventionally tilled winter wheat. Based on a snow storage routing developed by Steppuhn and McConkey (1988), the snow surface storage for all parts of the hill was 0.075 m; no snow drifts until this storage is filled. Snow depth and density were measured using a snow tube at 6 to 12 locations per hill position (Table 1). Measurements were made on 12/26 and 12/29, 1990, and 1/1, 1/8, 1/11, 1/12, 1/14, and 3/6, 1991. Meteorological information was collected at the Palouse Conservation Field Station 3 km north-

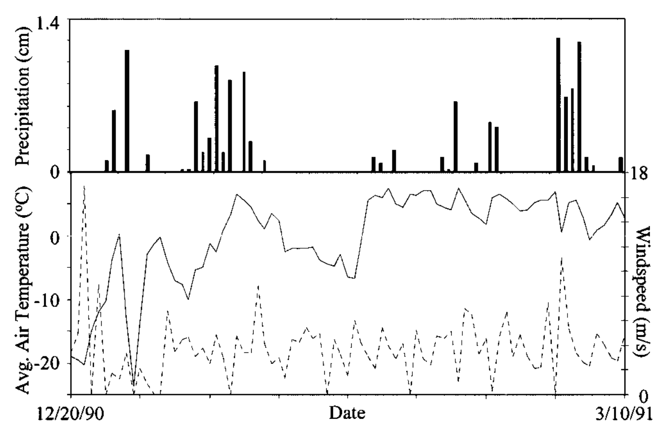

Fig. 2. Meteorological conditions at the Palouse Conservation Field Station during the simulation period. Bars=precipitation; solid line =average daily air temperature; and dashed line=windspeed.

west of Pullman, Wash. (Fig. 2). To assess the snowdrift model, the hill slope snowpack was simulated with and without the snowdrift model.

Not only was it impractical to measure snowdrift in the absence of snowmelt and accumulation, but these other processes impact snowdrift behavior. In particular, newly accumulated snow is typically less dense than old snow, and therefore more susceptible to drift. Also, refreezing of meltwater that previously percolated into the old, underlying snow will increase snow density, making this snow increasingly resistant to drift. It should also be noted that topography may influence snowmelt in ways that lead to snowdrift-like spatial patterns of snowcover. Although snow accumulation can be complicated, our model simply accumulates snow when precipitation falls at air temperatures below freezing. The density of the new snow, ρ_{ns} (kg m^{-3}), is a function of temperature as represented by (Goodison et al. 1981)

$$\rho_{ns}=50+3.4(T_a+15), \quad T_a<-15°\text{C}; \qquad \rho_{ns}=50, \quad T_a \geqslant -15°\text{C} \tag{12}$$

where T_a=the average daily air temperature (°C). For this study, no correction was made to precipitation gauge readings to account for possible inaccuracies in snowfall measurements. Snowmelt was simulated using an energy balance accounting for fluxes of radiation, evaporative heat exchange, sensible heat exchange, precipitation heat, and ground conduction. Although energy budgets like this have been developed by many researchers, this model is unique because the required meteorological parameters are maximum and minimum daily temperatures, windspeed, and precipitation; all other parameters can be reasonably estimated from these. The snow routines are fully described by Walter (1995) and Walter et al. (submitted for publication), and the most important differences between this model and others are the calculations for solar transmissivity, cloud cover for calculating atmospheric long-wave radiation, air-vapor density for calculating evaporative heat exchange, and surface albedo. Solar transmissivity is calculated as a function of the difference between maximum and minimum daily air temperatures using the Bristow-Campbell equation (Bristow and Campbell 1984) as modified by Ndlovu (1994). Cloud cover is a parameter in determining the atmospheric emissivity (Unsworth and Monteith 1975). Several methods were investigated for approximating this value, including relating cloud cover to solar transmissivity and simply assuming cloud cover was zero for less than trace precipitation and 100% for greater precipitation. There were negligible differences in snowmelt predictions among these methods, so the simplest was adopted. Air-vapor density was approximated by assuming that the minimum daily air temperature is equal to the dew-point (Campbell 1977) and used in the expression presented by Jensen et al. (1990) for saturated vapor density. An albedo function was developed using the relationships developed by the USACE (1960) and data presented by O'Neil and Gray (1973) for shallow snowpacks. The reliability of these approximations in simulating snowmelt is presented by Walter et al. (submitted for publication). Topographic effects on solar radiation were also simulated.

The snowmelt model alone was tested against measured data from four sites across the northern conterminous U.S.: Danville, Vt. (latitude~44° N) (Anderson and Whipkey 1977), Bloomville, N.Y. (latitude~41° N), Easton, Minn. (latitude~44° N) (Brooks 1997), and Troy, Ind. (latitude~47° N). Full results and a discussion are given by Water et al. (submitted for publication). At all four sites drift was minimal, the topography was level (although slight slopes were accounted for where noted), and the landscape was open, i.e., unforested. The sites and data collection methods are described fully by Walter et al. (submitted for publication). Because the focus of this paper is the snowdrift model, only the essential corroboration of the model with the melt/accumulation data is presented here.

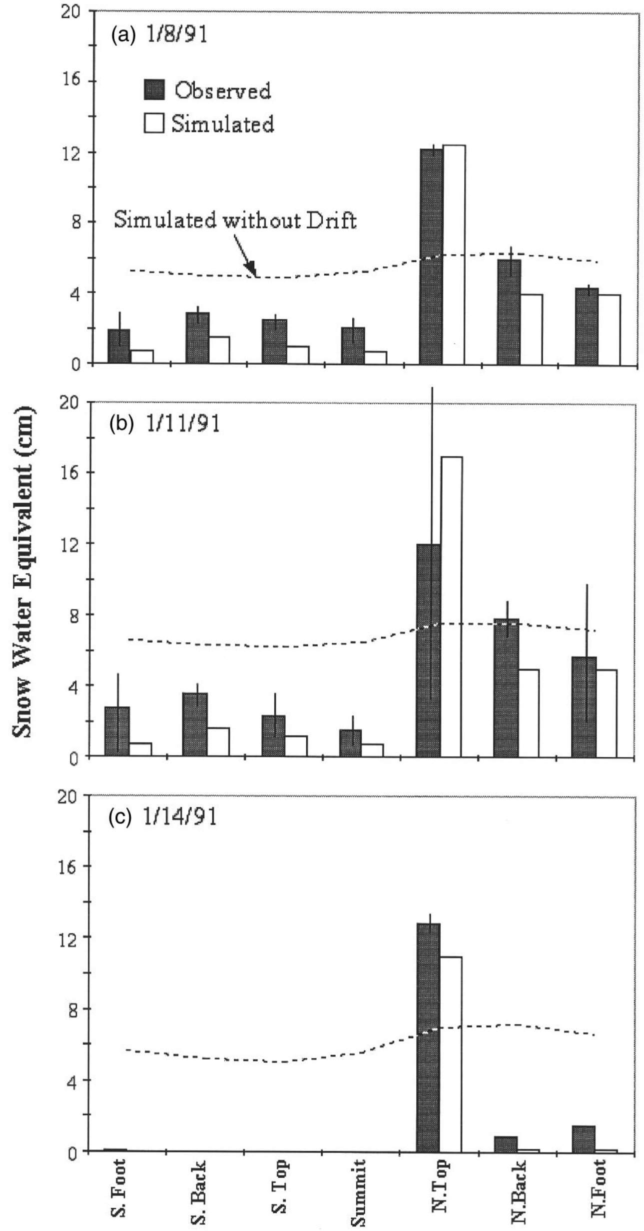

Fig. 3. Simulated (open bars) and observed (solid bars) snow water equivalent across the hill for (a) 1/8/91; (b) 1/8/91; and (c) 1/14/91. The dashed line shows the simulation results without snowdrift. Vertical lines are ±one standard deviation of measurements.

Results and Discussion

Fig. 3 summarizes the eastern Washington snowdrift results comparing the snowdrift model with observed data and no-drift snow model results (dashed-line in Fig. 3). Table 2 shows the statistical comparisons for each hill-slope position using all eight sampling dates. The snowdrift model's prediction of the overall observed spatial and temporal trends across the hill slope, i.e., using all data lumped together, is very good ($R^2=0.95$), especially compared

Table 2. Statistical Analysis for Snowdrift Model using Data from Pullman, Wash. Data Are Snow Water Equivalents (cm)

Slope aspect	Position[a]	Observed mean[b]	Simulated mean	Standard error	R^2	RMSE[c]	Relative difference[d]	Standard deviation[e]
South	Foot	1.5	0.6	0.22	0.32	1.4	0.92	0.3
	Back	1.7	1.0	0.56	0.77	1.1	0.64	0.3
	Top	1.6	0.6	0.10	0.64	1.4	0.89	0.2
	Summit	1.3	0.6	0.37	0.40	1.3	0.99	0.3
North	Top	10.3	9.7	0.56	0.75	3.3	0.32	3.7
	Back	3.7	2.1	0.30	0.94	2.0	0.54	0.4
	Foot	2.7	2.2	1.19	0.84	0.8	0.31	0.7

[a]Hillside positions correspond to those characterized in Table 1.

[b]Based on eight sampling dates and 6–12 measurements per hillslope position. Examples of standard deviations on specific days are presented in Fig. 3.

[c]Root Mean Square Error (RMSE): $\sqrt{\Sigma(\mathrm{Obs}-\mathrm{Sim})^2/N}$ where N is the sample size.

[d]Relative Difference: RMSE/$\overline{\mathrm{Obs}}$.

[e]Standard deviation among Obs.

to the no-drift model ($R^2=0.33$). This was surprising considering the simplicity with which this model adjusts windspeed and simulates the snow processes involved. Most notably is the observed and simulated development and persistence of a snowdrift near the hilltop on the north side of the hill. Similar trends are expected simply based on differences in solar insolation across the different hill slopes and aspects, but this modeled variability (dashed line in Fig. 3) did not account for the full range of observed variability (Fig. 3).

The snowdrift model corroboration at some hill-slope positions was better than at others, and at some locations the agreement was modest (Table 2); at all locations, the absolute errors were relatively small. At all locations, the snowdrift model performed better than the no-drift model (Fig. 3); for the no-drift model, RMSE=3.2–5.5 cm, relative difference (R.D.)=0.45–4.02, and $R^2=0.002$–0.31. Comparing the overall trends ($R^2=0.95$) to the correlations at specific points along the hill (Table 2), highlights the on-going debate among hydrological modelers about concerns that using integrated data to corroborate distributed models often overstates model performance.

Because some of the error observed in the snowdrift results (Fig. 3; Table 2) is partially due to errors in snowmelt and snow accumulation, we evaluated the melt/accumulation process independently. Fig. 4 and Table 3 summarize the results from the snowmelt-accumulation analysis for four sites in the northern conterminous U.S. There was generally good agreement between predicted and observed snow water equivalent (SWE) at all four sites [Fig. 4(a); Table 3]. Figure 4(b) shows an example of temporal trends in predicted and observed SWE for one site, Danville, Vt., in the absence of snowdrift. The slightly enhanced rate of snowmelt at the end of the March is due to an under prediction in albedo and, thus, enhanced melt due to solar radiation; solar radiation becomes an increasingly dominant part of the energy balance throughout March in Vermont. To assess the model's snowmelt performance in a more precise way, we compared simulated and observed total meltwater draining from the snowpack [Fig. 4(c)]; note that these data were only available at the Danville, Vt. site. The predicted snowmelt generally agreed with observed melt [Fig. 4(c)], although the comparison was poorer than the other snowmelt-accumulation results in part because of the extra complications of determining snow water holding capacity (Walter et al., submitted for publication). For comparison purposes, meltwater was also predicted with the well known U.S. Army Corps of Engineers (1960) equations using measured meteorological data (Anderson and Whipkey 1977). The U.S. Army Corps of Engineers (1960) model resulted in meltwater statistics of $R^2=0.36$ and standard error (STE)=0.75 cm, which suggests substantially worse performance than our melt model; we concluded that our model performed at least as well as other widely used snowmelt-accumulation models. It was not practical to evaluate snow accumulation independently, but Fig. 4(b) suggests that the model slightly over-predicted snow accumulation. The model also tended to slightly under-predict new snow density (data not shown). In short, the melt and accumulation components of the model used in this paper predicted SWE on the order of ~1 cm.

Discrepancies between measured and observed snowdrift results (Fig. 3; Table 2) were of the same magnitude as the errors (e.g., STE) for snowmelt, so it is difficult to evaluate the snowdrift errors independently. Furthermore, because of the interdependence of the processes involved, it was not possible to accurately isolate specific processes that contributed to errors. However, as expected, the snowdrift results (Table 2) were somewhat poorer than the snowmelt results (Table 3) because of accumulated errors. It appears that the drift model consistently overpredicted snow removal on the south face of the hill and somewhat over predicted snow deposition at the top of the north side of the hill [Fig. 3(b)]. The model received no explicit calibration because it was unclear what parameters could be meaningfully calibrated. For example, model results can be improved by adjusting and varying surface snow storage, but a rationale independent of direct calibration was not apparent. Nevertheless, it is interesting that the model correctly simulated the relative distribution of snow across the hill, even on the south side where the model corroboration was the worst. For example, the south-back side of the hill consistently had more snow than other locations on the south side of the hill, and this trend is duplicated in the model. The snow settling and densification portion of the model is probably the most suspect because it received no independent validation. One notable unknown is enhanced densification due to wind.

Note that the especially large difference between the no-drift and drift models over most of the hill slope suggests that sublimation as well as snow removal by wind may be an important process controlling snow distribution in these areas. This is consistent with sublimation field studies (Pomeroery and Li 2000). It is unclear, however, that simply assuming sublimation losses equal drift losses are a sufficient representation of the mechanism.

The north side of the hill was simulated substantially better than the south. This was in part due to deeper snowpacks for

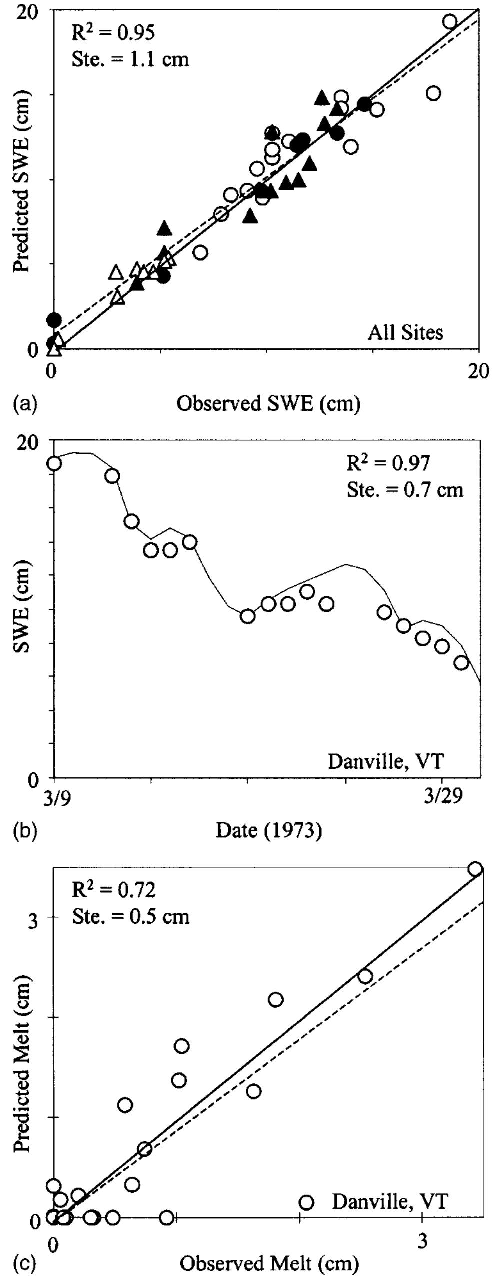

Fig. 4. Corroboration of the snowmelt model's simulated results with observed data. (a) Predicted versus observed snow water equivalent (SWE) for all four sites, open circles=Danville, Vt., solid circles =Bloomville, N.Y., open triangles=Easton, Minn., solid triangles =Troy, Id., solid line=1:1, dashed line=regression; (b) Example of temporal predicted (line) and observed (open circles) SWE trends at Danville, Vt.; (c) Predicted and observed meltwater at Danville, Vt., solid line=1:1, dashed line=regression.

which albedo estimates are more reliable. For very shallow snowpacks, as consistently observed on the south slope, the underlying ground albedo influences the surface albedo in ways that are difficult to predict (O'Neil and Gray 1973). This is consistent with the slightly early, simulated snow disappearance at the back and foot of the north side [Fig. 3(c)].

It is probable that the results presented here are close to the worst-case for this model because the snowpack is shallow and transient, and winddrift was substantial. The difficulties with modeling shallow snowpacks were briefly mentioned earlier in this paper and have been noted by other researchers (O'Neil and Gray 1973, Flerchinger and Saxton 1989). In light of this and the relative simplicity with which the snowdrift processes were simulated, this snowdrift model performs well. The STEs for the snowdrift model (Table 2) were similar to those for the snowmelt model (Table 3); if this test represents a worst case scenario for the snowdrift model, it might be expected that the errors would be similar for deeper, more persistent snowpacks.

A simple sensitivity test was performed by adjusting the components of the model that rely on the most unsubstantiated assumptions by $\pm 10\%$ of the values used in this analysis and reanalyzing the results. Specifically tested were the assumptions that sublimation is equal to drift losses, snow densification rates [Eqs. (9)–(11)], the calculated radiation fluxes, and local wind adjustments [Eqs. (5)–(7)]. No single parameter adjustment changed the model results by more than 10%. As expected, decreasing sublimation rates increased SWE on the south side of the hill by about 5–10% and did not change SWE on the north side of the hill substantially, which improved the overall results. As with surface snow storage discussed earlier, sublimation could be calibrated to give good results for the south side of the hill. The model was also run using a single soil layer with average density, and results were similar to the "no-drift" results shown in Fig. 3. Thus, the most important parameter appears to be the density of new snowfall.

The observations from this simple study can be used to apply this model to a very complicated landscape. Because substantial drifting appears to be isolated to very small regions within the landscape, a suggested approach is to identify windward edges of vegetated areas and lee sides of hills, snow fences, etc. and divide the region's total transported snow among these areas. Evaluating the performance of various methods for partitioning the snowdrift is the next logical phase of this research.

Conclusion

This snowdrift model performed well over a single hillside in eastern Washington, especially relative to a similar model that did not account for snowdrift (Fig. 3). The model performance evaluated across the hill and simulation period ($R^2 = 0.95$) was better than the performance evaluated at individual points along the hill (Table 2; $R^2 = 0.32–0.94$). Standard errors were low, generally <1 cm, which was similar to errors in modeling snowmelt in the absence of snowdrift. The snowdrift model captured both snow removal and snow deposition processes, although it slightly over predicted snow removal, probably due to over predictions in sublimation. The model's successful performance is especially noteworthy given its simplicity relative to much more complicated models, most of which do not account for snow drifting.

The next phase of this research is to evaluate simple ways to distribute snowdrift across a complicated landscape. This work, like many aspects of distributed hydrological modeling, hinges on the development of better ways to evaluate distributed processes and parameters. Additional work focusing on snow metamorphosis under windy conditions could also improve this model's reliability.

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Table 3. Statistical Analysis for the Snowmelt Model (Walter et al. (submitted for publication) for Complete Analysis)

Location	Observed mean	Simulated mean	Standard error	R^2	Standard deviation	Simulate period (N^a)
Bloomville, N.Y.						1/25–3/7/97
SWE (cm)	8.2	8.4	0.8	0.98	5.8	(8)
Easton, Minn.						12/28/94–3/2/95
SWE (cm)	2.9	3.3	0.6	0.94	2.1	(10)
Troy, Id.						12/10/99–4/1/00
SWE (cm)	9.0	9.5	1.4	0.89	4.1	(13)
Danville, Vt.						3/9–3/31/73
SWE (cm)	14.4	14.4	0.6	0.97	2.0	(16)
Snow depth (cm)	39.4	44.9	3.6	0.83	5.3	
Melt water ($cm \cdot day^{-1}$)	0.79	0.70	0.50	0.72	0.93	

Note: SWE=snow water equivalent.

[a]Number of measurements.

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